

Hydrological application of remote sensing: Surface states – Snow

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Popular Summary

Remote sensing research of snow cover has been accomplished for nearly 40 years. The use of visible, near-infrared, active and passive-microwave remote sensing for the analysis of snow cover is reviewed with an emphasis on the work on the last decade.

Statement of Significance

Satellite remote sensing has been used to map snow cover for nearly 40 years. Decade-scale climate-data-record-quality records of snow-covered area are already in existence for the Northern Hemisphere plus are useful in climate models, and are invaluable for other climate studies. We can now extend the snow-covered area record using SMMR, SSM/I MODIS and AMSR data, to the global scale, and provide climate-data-quality maps of not only snow-covered area, but also SWE and snow albedo using visible, near infrared, and passive-microwave sensors such as the MODIS and AMSR. Production of these maps is necessary in order to analyze long-term climate trends. Whether the snow products are used for initial conditioning or as a forcing variable in a model, or whether the products are used in their own right, the role of remotely sensed observations of snow will continue to be important and are set to play an increasingly important role in climate and hydrological forecasting.

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Introduction

Snow is a porous, permeable aggregate of ice grains (**Bader, 1962**). Snow crystals nucleate and begin their growth in clouds where the temperature is below 0° C. Snow crystals can occur in a variety of relatively flat hexagonal or six-sided shapes (Figure 1). However, they can also occur as elongated columns and needles. Differences in snow crystals are a result of variations in the temperature and humidity of the atmosphere at the time of their formation and the action of the wind during their decent to the ground (**Male, 1980**). As a result, there is a myriad of possible shapes that can form depending primarily on the cloud temperature at the time the water vapor freezes.

Freshly-fallen snow almost immediately begins to compact and metamorphose, initially preserving the original shape of the snow crystal. The constant jostling and rubbing of crystals against each other causes protuberances to become chipped and broken. As the snow settles under its own weight, melts and refreezes and is buffeted by the wind, individual crystals are further altered so after a few days they have little resemblance to their original shape.

In the absence of a temperature gradient in the snowpack, snowflakes undergo “destructive metamorphism,” and become more rounded over time, with typically slower growth rates being more characteristic of the more-rounded crystals. “Constructive metamorphism,” when large grains grow at the expense of small grains, occurs when there is a thermal and vapor gradient in the snowpack and snow grains at the base of the pack grow at the expense of smaller grains hence the crystals develop distinctive shapes (**Colbeck, 1982**). Faster growth rates give rise to the faceted crystals. Dry snow will

metamorphose into large depth-hoar grains when subjected to a strong temperature gradient, and grain growth and dry-snow metamorphism control the movement and redistribution of mass, chemical species and isotopes in the snowpack (**Sturm and Benson, 1997**).

Seasonal snowpacks develop from a series of winter storms and are often created by various forms of precipitation such as rain and freezing rain; diurnal melting and refreezing and wind action are also important in the ultimate development of a snowpack. Thus seasonal snow cover usually develops a layered structure in which ice layers alternate with layers of a coarser texture (**Male, 1980**).

The geographical extent of snow cover over the Northern Hemisphere varies from a maximum of $\sim 46 \times 10^6 \text{ km}^2$ in January and February, to a minimum of $\sim 4 \times 10^6 \text{ km}^2$ in August; between 60 – 65% of winter snow cover is found over Eurasia, and most mid-summer snow cover is in Greenland (**Frei and Robinson, 1999**). Numerous studies have shown the importance of accurate measurements of snow and ice parameters as they relate to the Earth's climate and climate change (for example, see **Martinelli, 1979**; **Dewey and Heim, 1981 & 1983**; **Barry, 1983, 1984 & 1990**; **Dozier, 1987**; **Ledley et al., 1999**; **Foster et al., 1987 & 1996**; **Serreze et al., 2000**). Measurements have become increasingly sophisticated over time. And in addition, as the length of the satellite record increases, it becomes easier to determine trends that have climatic importance.

Satellite remote sensing technology virtually revolutionized the study of snow cover. Satellites are well suited to the measurement of snow cover because the high albedo of snow presents a good contrast with most other natural surfaces except clouds. Weekly snow mapping of the Northern Hemisphere using National Oceanographic and Atmospheric Administration (NOAA) data began in 1966 and continues today in the United States, but with improved resolution and on a daily basis (**Matson et al., 1986**; **Ramsay, 1998**). And in addition, using Earth Observing System (EOS) Moderate Resolution Imaging Spectroradiometer (MODIS) data, beginning in early 2000, global snow cover has been mapped on a daily basis at a spatial resolution of up to 500 m (**Hall et al., 2002a**).

In addition to the visible/near-infrared data, both passive and active microwave data have been useful for mapping snow and determining snow wetness and snow-water equivalent (SWE) (**Ulaby and Stiles, 1980**) since the early 1970s. Using passive-microwave data, snow extent and SWE may be determined globally on a daily basis since the launch of the Nimbus-7 Scanning Multichannel Microwave Radiometer (SMMR) (**Chang et al., 1987**), and continuing with the May 2002 launch of the Advanced Microwave Scanning Radiometer (AMSR) (**Kelly et al., 2003**).

Three of the most important properties of snow cover are depth, density and water equivalent (**Pomeroy and Gray, 1995**). If the snow depth and density are known, then the snow-water equivalent (SWE) may be calculated. SWE is a hydrologically-important parameter as it determines the amount of water that will be available from snowmelt.

In this chapter we will show how remote sensing is used to study snow-covered area, SWE, snow wetness and snow albedo.

Snow Mapping using Visible and Near-Infrared Sensors

Early attempts to forecast runoff from the areal extent of snow cover used terrestrial photographs (**Potts, 1937**). Other observations from aircraft can be made, i.e. locating and mapping snow-cover extent and the location of the snowline. Along with volume, the areal extent of snow cover has been used to predict snowmelt runoff and to forecast floods.

Because of its high albedo, snow was easily observed in the first image obtained from the Television Infrared Operational Satellite-1 (TIROS-1) weather satellite in 1960. Data from meteorological satellites and manned spacecraft were useful in determining snowline elevation, delineating snow boundaries, and observing changes in snow conditions due to rising temperatures and rain-on-snow events (**Singer and Popham, 1963**). Data from Environmental Science Services Administration (ESSA) operational satellites were used as early as the mid-to-late 1960s to determine areal extent of snow cover.

A major step forward in snow mapping came with the advent of the Landsat series of sensors beginning in 1972. Landsat-1 carried a Multispectral Scanner (MSS) sensor with 80-m spatial resolution. With Landsat data came the ability to create detailed basin-scale snow-cover maps on a regular basis when cloudcover permitted. At first, the repeat-pass interval for the Landsat satellite was 18 days, but this was decreased to 16 days with the launch of Landsat-4 in 1982. Landsats-4 and -5 carried a Thematic Mapper (TM) sensor with 30-m resolution, and Landsat-7 carries an Enhanced Thematic Mapper Plus (ETM+) with spatial resolution of 30 m except in the panchromatic band where the resolution is 15 m. (Landsat-6 was lost after it failed to reach orbit in 1993.) Though the Landsat series has provided high-quality, scene-based snow maps, the 16- or 18-day repeat-pass interval of the Landsat satellites is not adequate for most snow-mapping requirements, especially during spring snowmelt.

Operational snow-cover mapping in the United States. The Satellite Analysis Branch of NOAA's National Environmental Satellite, Data, and Information Service (NESDIS) began to generate Northern Hemisphere Weekly Snow and Ice Cover analysis charts derived from NOAA's Geostationary Operational Environmental Satellite (GOES) and Polar Orbiting Environmental Satellite (POES) visible satellite imagers in November 1966. Maps were manually constructed and the spatial resolution of the charts was 190 km. Since 1997 the Interactive Multi-Sensor Snow and Ice Mapping System (IMS) is used by analysts to produce products daily at a spatial resolution of about 25 km, and utilizes a variety of satellite data to generate the maps (**Ramsay, 1998**). This snow-cover record has been studied carefully (**Robinson et al., 1993; Robinson, 1997 & 1999**) and has been reconstructed following adjustment for inconsistencies that were discovered in the early part of the data set (**Robinson and Frei, 2000; Frei et al., 1999**). Results show

that the Northern Hemisphere annual snow-covered area has decreased (**Robinson et al., 1993; Brown and Goodison, 1996; Hughes and Robinson, 1996; Hughes et al., 1996; Armstrong and Brodzik, 1998 & 2001; Frei et al., 1999; Brown, 2000**); satellite data show a decrease of about 0.2% per year from 1978 - 1999 (**Armstrong and Brodzik, 2001**).

The National Operational Hydrologic Remote Sensing Center (NOHRSC) snow-cover maps, generated by National Weather Service NOHRSC hydrologists are distributed electronically in near real time, to local, state and federal users during the snow season (**Carroll, 1987 & 1995; Carroll et al., 2001**). The NOHRSC 1-km maps are generated primarily from the NOAA polar-orbiting satellites and GOES satellites to develop daily digital maps depicting the areal extent of snow cover for the coterminous United States, and Alaska, and parts of southern Canada.

Other snow maps. Landsat data have been used for measurement of snow-covered area over drainage basins (**Rango and Martinec, 1979 & 1982**). The Landsat TM and ETM+ have been especially useful for measuring snow cover because of the short-wave infrared band – TM band 6 (1.6 μm) which allows snow/cloud discrimination. The reflectance of snow is low and the reflectance of most clouds remains high in that part of the spectrum. Various techniques, ranging from visual interpretation, multispectral image classification, decision trees, change detection and ratios (**Kyle et al., 1978; Bunting and d'Entremont, 1982; Crane and Anderson, 1984; Dozier, 1989; Romanov et al., 2000; Hall et al., 2002a; Romanov and Tarpley, 2003**) have been used to map snow cover. Other spectral and threshold tests are also used.

The Moderate Resolution Imaging Spectroradiometer (MODIS) was first launched in December 1999 on the Terra spacecraft. MODIS data are now being used to produce daily and 8-day composite (Figure 2) snow-cover products from automated algorithms at Goddard Space Flight Center in Greenbelt, Maryland (**Hall et al., 2002a**). The products are transferred to the National Snow and Ice Data Center (NSIDC) in Boulder, Colorado, where they are archived and distributed via the EOS Data Gateway (EDG) (**Scharfen et al., 2000**). The Aqua satellite was launched in 2002 with a second MODIS instrument that will enable snow-covered area measurements to be extended farther into the future.

The MODIS maps provide global, daily coverage at 500-m resolution, and the climate-modeling grid (CMG) maps are available at 0.05° resolution which is ~5.6-km at the Equator. The CMG map, designed for use by climate modelers, provides a global view of the Earth's snow cover in a geographical projection with fractional snow cover reported in each cell. The automated MODIS snow-mapping algorithm uses at-satellite reflectances in MODIS bands 4 (0.545-0.565 μm) and 6 (1.628-1.652 μm) to calculate the normalized difference snow index (NDSI) (**Hall et al., 2002a**). Other threshold tests are also used including use of the Normalized Difference Vegetation Index (NDVI) together with the NDSI to improve snow mapping in forests (**Klein et al., 1998**). **Bussi res et al. (2002)** compared MODIS-derived snow maps with SSM/I-derived snow maps and found generally good correspondence. **Mauer et al. (2003)** compared MODIS and NOHRSC data for the Columbia River Basin, U.S.A., and found that the maps were

comparable, but the MODIS maps generally provided more cloud-free data than did the NOHRSC maps. Add more from Bussi eres et al. (2002), Klein and Barnett (2003) and Simic et al. (HP in press) – get information from our MODIS chapter of John Qu’s book.

In the near future, percent snow cover or fractional snow cover in each pixel will be provided along with daily snow albedo (Klein and Stroeve, 2002) in the 500-m products (Salomonson and Appel, submitted). Because cloudcover often precludes the acquisition of snow cover from visible and near-infrared sensors, the daily maps are composited, and 8-day composite maps as well as daily maps are available.

Fractional snow cover or subpixel snow mapping. Fractional snow cover (FSC) utilizing Landsat and MODIS data has been calculated to exploit sub-pixel information. Much of this work has relied on spectral-mixture modeling (see Nolin et al., 1993; Rosenthal and Dozier, 1996; Vikhamer and Solberg, 2002), and neural networks (Simpson and McIntire, 2001) but does not provide global coverage. Painter et al. (2003) developed an automated model that couples spectral-mixture analysis with a radiative transfer model to retrieve subpixel snow-covered area and effective grain size from Airborne Visible/Infrared Imaging Spectrometer (AVIRIS) data. Other work using MODIS data (Kaufman et al., 2002) has developed algorithms to map FSC globally. Recently, Salomonson and Appel (submitted) have extended the use of the NDSI to provide FSC globally with absolute errors of 0.1 or less over the whole range of FSC from 0 to 100%.

In Norway, the Norwegian Water Resources and Energy Directorate (NVE) and the Troms  Satellite Station (TSS) produce snow maps from Advanced Very High Resolution Radiometer (AVHRR) data using band 2 (0.7-1.1 μm) for snow mapping, and bands 3 (3.6 - 3.9 μm), 4 (10.3-11.3 μm) and 5 (11.5 - 12.5 μm) for snow/cloud discrimination (Anderson, 1982; K nig et al., 2001). An upper limit of 100% snow covered is determined from a glacier or 100% snow-covered region and a lower limit of 0% is determined from water or land areas, and percentage of snow cover is interpolated linearly thus deriving FSC (K nig et al., 2001).

Albedo. The spectral albedo of a surface is the upflux divided by the downflux at a particular wavelength (Warren, 1982). The broadband albedo is the reflectance across the reflective part of the solar spectrum. The spectral albedo of fresh snow in the visible region of the spectrum remains high but decreases slowly as snow ages, but in the near-infrared the spectral albedo of aging snow decreases considerably as compared to fresh snow (O’Brien and Munis, 1975; Warren and Wiscombe, 1980). Maxima in the spectral albedo values across the spectrum are found at 1.1, 1.3, 1.8 and 2.2 μm , and they correspond to local minima in the absorption coefficient of ice (Wiscombe and Warren, 1980).

Broadband albedo decreases when grain size increases as the snow ages (Choudhury and Chang, 1979) and melting causes snow grains to grow and bond into clusters (Dozier et al., 1981; Grenfell et al., 1981; Warren, 1982). Snow albedo may decrease

by >25% within just a few days as grain growth proceeds (**Nolin and Liang, 2000**). For example, **Gerland et al. (1999)** measured a maximum albedo >90% on Svalbard, Norway, before melt onset, and ~60% after melt had progressed in the spring when the snow was considered old, but still clean. Grain size may be estimated using remotely-sensed data (**Dozier, 1984; Nolin and Dozier, 1993**). The albedo of a snow cover is also influenced by the albedo of the land cover that it overlies.

With the onset of surface melting and associated grain size increases, the near-infrared reflectance decreases dramatically (**Warren, 1982**) (Figure 3). The near-infrared albedo of snow is very sensitive to snow grain size while visible albedo is less sensitive to grain size, but is affected by snow impurities. Modeling by **Warren and Wiscombe (1980)** demonstrates that small, but highly absorbing particles can lower the snow albedo in the visible part of the spectrum by 5-15% compared to pure snow.

Though the reflectance of freshly-fallen snow is nearly isotropic (**Dirmhirn and Eaton, 1975**), as snow ages the specular reflection component increases especially in the forward direction and with solar zenith angle (SZA) (**Salomonson and Marlatt, 1968**), and the anisotropic nature of snow reflectance increases with increasing grain size (**Steffen, 1987**). Effective snow grain radii typically range in size from ~50 μm for new snow, to 1 mm for wet snow consisting of clusters of ice grains (**Warren, 1982**).

Snow albedo increases at all wavelengths with the SZA. Additionally, cloudcover normally causes an increase in spectrally-integrated snow albedo due to multiple reflections caused by clouds (**Grenfell and Maykut, 1977; Warren, 1982**).

Early attempts to measure snow albedo remotely were conducted from aircraft (**Bauer and Dutton, 1962; Hanson and Viebrock, 1964; McFadden and Ragotzkie, 1967; Salomonson and Marlatt, 1968; Dirmhirn and Eaton, 1975**). More recently, however, detailed field, aircraft and satellite studies have been undertaken to derive quantitative measurements of snow reflectance and albedo (for example, see **Steffen, 1987; Hall et al., 1989; Duguay and LeDrew, 1992; Winther, 1993; Knap and Oerlemans, 1996; Stroeve et al., 1997; Winther et al., 1999; Greuell et al., 2002**).

Some researchers have measured the albedo of snow-covered lands using satellite data on a hemispheric scale (e.g., **Kukla and Robinson, 1980; Robock, 1980; Robinson and Kukla, 1985; Robinson, 1993**). Both **Robinson et al. (1986)** and **Scharfen et al. (1987)** constructed basin-wide albedo maps and observed differences in the timing of the melt between years. **Robinson and Kukla (1985)** used Defense Meteorological Satellite Program (DMSP) imagery (spectral range – 0.4 - 1.1 μm) to derive a linear relationship between the brightest snow-covered arctic tundra and the darkest snow-covered forest which were assigned albedos of 0.80 and 0.18, respectively. Scene brightness was then converted to surface albedo by linear interpolation. The surface brightness is a function of the type and density of vegetation and the depth and age of snow (**Robinson and Kukla, 1985**). The derived “maximum” surface albedo values were useful for climate modeling (**Ross and Walsh, 1987**).

Surface albedo has also been derived from Landsat MSS and TM data. One approach, based on exact solutions of the radiative-transfer equation for upwelling intensity, requires known albedo values derived in each Landsat scene at different points (**Mekler and Joseph, 1983**). Other approaches rely on a narrowband to broadband conversion to derive albedo (**Brest and Goward, 1987; Hall et al., 1989; Duguay and LeDrew, 1990; Knap et al., 1998; Winther et al., 1999**). **Knap and Reijmer (1998)** used Landsat data to derive the Bi-directional Reflectance Distribution Function (BRDF) to describe the complete distribution of the anisotropic reflectance of snow, and **Greuell and de Ruyter de Wildt (1999)** used BRDF to correct for anisotropic reflectance. The BRDF is the physical property that determines the amount and angular distribution of reflected radiance from a surface (**Nicodemus et al., 1977; Nolin and Liang, 2000**).

AVHRR data have been used to map changes in albedo over the Greenland Ice Sheet during the spring and summer months (**Knap and Oerlemans, 1996; Nolin and Stroeve, 1997; Stroeve et al., 1997**). While **Stroeve et al. (1997)** found a good correspondence with satellite-derived and surface-measured albedo before snowmelt, after snowmelt began, melt-water ponding on the ice surface precluded accurate comparisons between the satellite-derived and surface-measured albedo.

Surface albedo over the Greenland ice sheet has also been measured using data acquired by the EOS Multi-angle Imaging SpectroRadiometer (MISR) instrument. MISR images the surface using nine discrete, fixed-angle cameras, one nadir-viewing and four viewing angles in forward and aftward directions along the spacecraft track. Compared with in-situ measurements at five different sites, the surface albedo derived from two different methods using the MISR data showed good agreement (within about 6%) according to work done by **Stroeve and Nolin (2002)**. Although the atmosphere is relatively thin over the ice sheet, atmospheric attenuation is significant in the visible and near-infrared wavelengths, and the 6S radiative transfer model (**Tanre et al., 1992**) was used to derive the results.

A near-surface global algorithm has been developed to map snow albedo using MODIS data (**Klein and Stroeve, 2002**). In deriving albedo, atmospherically-corrected MODIS surface reflectances in individual MODIS bands for snow-covered pixels located in non-forested areas are adjusted for anisotropic scattering effects using a discrete ordinates radiative transfer model (DISORT) and snow optical properties (**Klein and Stroeve, 2002**). Currently in the algorithm, snow covered forests are considered to be Lambertian reflectors. The adjusted spectral albedos are then combined into a broadband albedo measurement using a narrow-to-broadband conversion scheme developed specifically for snow by Shunlin Liang (**written communication, 2003**) (**Liang, 2000; Klein and Hall, 2000; Klein and Stroeve, 2002**). Thus derived, a near-global snow albedo product (Figure 4) is available from February 2000 to the present through the National Snow and Ice Data Center in Boulder, CO.

Snow Mapping and SWE determination using Microwave Sensors

Introduction. In the microwave part of the spectrum (300-1 GHz, or 1 mm to 30 cm wavelength), remote sensing can be accomplished either by measuring emitted radiation with a radiometer or by measuring the intensity of the return (in decibels) of a signal sent by a radar. Microwave measurements have the capability to respond to the bulk properties of a snowpack as well as to variations in other surface and subsurface features because microwaves can penetrate the snowpack and thus provide information on snow depth and SWE when the snowpack is dry. Additionally, the microwave part of the spectrum allows remote observation of snow cover under nearly all weather and lighting conditions.

Theoretically, the dielectric constant of snow consists of the sum of a real and an imaginary part. Snow is a mixture of air and ice, the dielectric constant of air being 1.0 and ice 3.17 \pm 0.07 for frequencies from 1 MHz to well above the microwave region (Evans, 1965). Snow has a dielectric constant between 1.2 and 2.0 when the snow densities range from 0.1 to 0.5 g/cm³ (Hallikainen and Ulaby, 1986).

The dielectric properties of snow at a given microwave frequency are generally dependent on the relative proportion of liquid and solid water in the snow by volume. Even at temperatures $<0^{\circ}\text{C}$, liquid-like water exists in thin films surrounding, and bound to ice crystals (Hobbs, 1974), but is considered to be dry since it contains no "free" liquid water (Leconte et al., 1990). Snow that contains a large amount of liquid water ($>5\%$ by volume) has a high dielectric constant (>35 below 20 GHz) relative to that of dry snow.

Microwave emission from a layer of snow over a ground medium consists of contributions from the snow itself and from the underlying ground. Both contributions are governed by the transmission and reflection properties of the air-snow and snow-ground boundaries and by the absorption/emission and scattering properties of the snow layer.

If a dry snowpack contains ice and snow layers, specular reflection may occur resulting in strongly-enhanced backscatter (Mätzler and Schanda, 1984). Or, if the grain sizes of a dry snowpack are large enough relative to the microwave wavelength, volume scattering will occur. Otherwise, the signal is returned mainly from the ground/snow interface.

Therefore, longer wavelengths, in general, travel almost unaffected through dry snow. X-band (2.4-3.75 cm, 8.0-12.5 GHz) or lower frequencies (longer wavelengths) are not generally useful for detecting and mapping thin, dry snow because the size of snow particles is much smaller than the size of the wavelength. Thus there is little chance for a microwave signal to be attenuated and scattered by the relatively small ice crystals comprising a snowpack (Waite and MacDonald, 1970; Ulaby and Stiles, 1980 & 1981). Wavelengths longer than $\sim 10 - 15$ cm travel unaffected through most dry seasonal snowpacks (Bernier, 1987).

For snow crystals of a radius $> \sim 0.1$ mm (a seasonal snowpack might have grain sizes ranging from 0.1 to 0.5 mm through the season), scattering dominates emission at higher

(>15 GHz) microwave frequencies (Ulaby et al., 1986). Absorption is determined primarily by the imaginary part of the refractive index. In dry snow, the imaginary part is very small, several orders of magnitude smaller than for water (Ulaby and Stiles, 1980).

Volume scattering increases with snow grain size and internal layering and with an increase in the amount of snow. Radiation at wavelengths comparable in size to the snow crystal size (about 0.05-3.0 mm) is scattered in a dry snowpack according to Mie scattering theory. There currently are no operational radars operating at these wavelengths, but there are passive-microwave sensors that operate at those wavelengths.

Passive-Microwave Snow Mapping and SWE Determination. In the microwave part of the spectrum, the radiation emitted from a perfect emitter is proportional to its physical temperature, T . However, most real objects emit only a fraction of the radiation that a perfect emitter would emit at its physical temperature. The equivalent temperature of the microwave radiation thermally emitted by an object is called its brightness temperature, T_B , expressed in Kelvins. This fraction defines the emissivity, E , of an object (Chang et al., 1976). In the microwave region,

$$E = T_B/T \quad [1]$$

Microwave emission from a layer of snow over a ground medium consists of two contributions: 1) emission by the snow volume, and 2) emission by the underlying ground. Both contributions are governed by the transmission and reflection properties of the air-snow and snow-ground interfaces and by the absorption/emission and scattering properties of the snow layer (Stiles et al., 1981), and a myriad of physical parameters that affects the emission (Derksen et al., 2002). As an electromagnetic wave emitted from the underlying surface propagates through a snowpack, it is scattered by the randomly-spaced snow particles in all directions. As the snowpack grows deeper, there is more loss of radiation due to scattering, and the emission of the snowpack is reduced, thus lowering the T_B . The deeper the snow the more crystals are available to scatter the upwelling microwave energy, and thus it is possible to estimate the depth and water equivalent of the snow.

Snow grains scatter the electromagnetic radiation incoherently and are assumed to be spherical and randomly spaced within the snowpack. Although snow particles are generally not spherical in shape, using Mie theory, their optical properties can be simulated as spheres (Chang et al., 1976). We will discuss the scattering power of a variety of different snow crystal shapes, and their effect on the microwave emission of a snowpack in the next section.

A wet snowpack radiates like a blackbody at the physical temperature of the snow layer, and is therefore indistinguishable from snow-free soil (Kunzi et al., 1982). The dielectric constants of water, ice and snow are different enough so that even a little surface melting causes a strong microwave response (Schanda et al., 1983; Foster et al.,

1987). The scattering loss decreases drastically with increasing liquid water content (free water) and becomes negligible for values above about 1% (Hallikainen, 1984).

For dry snow, the naturally-emitted microwave radiation of a snowpack is related to several physical properties of the snow. These properties include the number of snow grains along the emission path (the snow depth in cm), the size of grains (grain radius in mm) and the packing of the grains (volume fraction in % or density in kg m^{-3}). Such components control the propagation of radiation, especially at higher frequencies (e.g. 36 GHz) but affect the microwave response less at lower frequencies (e.g. 18 GHz). The brightness temperature difference between 18 GHz and 36 GHz ($T_B 18 - T_B 36$) is used to minimize the effect of snow temperature on the microwave emissivity. This is the principle that has been used to estimate SWE and snow depth from passive microwave instruments (Chang et al., 1976 & 1982; Kunzi et al., 1976 & 1982; Goodison and Walker, 1994; Goodison et al., 1986; Grody and Basist, 1996; Foster et al., 1997; Kelly and Chang, in press). Experiments and applications have shown that:

$$SD = a(T_B 18H - T_B 36H) \text{ [cm]}, \quad [2]$$

where SD is the snow depth in cm, a is constant of 1.59 for SMMR and is derived from radiative transfer experiments (Chang et al., 1982) and $T_B 18H$ and $T_B 36H$ are the horizontally polarized brightness temperatures measured by the spacecraft at 18 and 36 GHz respectively. If SWE is desired, a is set to a value of 4.8. Research into estimation of SWE and snow depth from passive microwave instruments has used this principle and an example of its application to AMSR-E data is shown in Figure 5 (Kelly et al., 2003).

Currently, SWE of a dry snowpack can be estimated with passive-microwave sensors such as the Special Sensor Microwave/Imager (SSM/I), and the Advanced Microwave Scanning Radiometer (AMSR) (Table 1), which was launched on the Aqua satellite in May of 2002. In Canada, SSM/I data are used to provide operational SWE map products (Figure 6) for the Canadian prairie region.

Forest cover can adversely affect the SWE retrieval accuracy by reducing the characteristic scattering response from snow by suppressing the $T_B 18H - T_B 36H$ signal (Tiuri and Hallikainen, 1981; Hall et al., 1982; Hallikainen, 1984; Kurvonen and Hallikainen, 1997). Foster et al. (1997) attempted to correct for this effect by incorporating forest fraction into equation (1) such that:

$$SD = a(T_B 18H - T_B 36H) / (1 - ff) \text{ [cm]} \quad [3]$$

where ff is the per pixel forest fraction (expressed as a unit percent) that ranges from 0% to 50% (fractions greater than 50% are set to 50%). This effect improved the retrieval accuracy in forested regions.

Snow-grain size is another important parameter that influences the microwave brightness temperature. A model was developed to study the growth of the depth-hoar layer at the base of the snowpack on the Arctic Coastal Plain of Alaska during the winter, and

compared to brightness temperature as derived from the Scanning Multichannel Microwave Radiometer (SMMR) by **Hall et al. (1986)**. Results showed that an approximately 20 K lower T_B was found at inland sites with a comparable snow depth, but a thicker depth-hoar layer than was present at coastal sites with a thinner depth-hoar layer. Thus it is necessary to characterize snow grain size on a regional basis to enhance the accuracy of snow retrievals using passive-microwave data (**Armstrong et al., 1993**). Using SSM/I data, **Mognard and Josberger (2002)** modeled seasonal changes in snow-grain size using a temperature-gradient approach. This information was then used to parameterize the retrieval of snow depth in the northern Great Plains during the 1996-1997 winter season. Taking this approach further, **Kelly et al. (2003)** have recently developed a methodology to estimate snow grain size and density as they evolve through the season using SSM/I and simple statistical growth models. These estimated variables are then coupled with a dense media radiative transfer model (DMRT), described in **Tsang and Kong (2001)** and **Chang et al. (2003)**, to estimate SWE.

As early as 1972, **Meier (1972)** was able to map the snowline on Mt. Rainier in Washington State in the U.S., using the 270 K T_B line and a single channel (19.35 GHz) of an airborne radiometer. Since then, many different algorithms to map snow cover and SWE using passive-microwave data have been developed and tested (e.g., **Rott et al., 1981; Kunzi et al., 1982; Chang et al., 1982; Goodison, 1989; Goodison and Walker, 1994; Mätzler, 1987; Hallikainen and Jolma, 1992; Grody and Basist, 1996 and Derksen et al., 2002; Pivot et al., 2002; Walker et al., 2002**). Some appear to work better under certain conditions than others, and it is now well accepted that there is no algorithm that is ideal globally. The basic algorithm must be tuned to individual land-cover types (**Walker and Goodison, 2000**). For example, **Walker and Goodison (1993)** developed a wet-snow indicator using the SSM/I 37 GHz polarization difference for the Canadian Prairies, and **Goita et al. (2003)** developed separate algorithms, both based on the vertically-polarized difference index using 18 and 37 GHz data from SSM/I to map SWE in deciduous and coniferous forest types, respectively.

Foster et al. (1997) used a forest factor related to albedo, and as a result, the passive-microwave-derived estimates of snow mass were closer to observed values in the boreal forest in North America and Eurasia than when the forest factor was not used. The differences between the day and night brightness temperatures indicate the presence of liquid water in the snowpack. Early in the snow season, the difference is small, indicating the absence of liquid water in the snowpack. As spring approaches the difference increases, indicating the presence of liquid water during the day, and then the pack re-freezes at night. When liquid water does not refreeze at night the difference again becomes small, and the snowpack is ripe and will soon begin to melt (**Josberger et al., 1993**). **Ramage and Isacks (2002 and 2003)** used SSM/I-derived diurnal-amplitude variations to detect early melt on snow-covered icefields in southeast Alaska, and found that melt timing correlates well with nearby stream hydrographs.

Derksen et al. (2003) produced a time series for central North America for the winter season from 1978 through 1999, using both SMMR and SSM/I data. They found evidence of systematic SWE underestimation during the SMMR years, and

overestimation of SWE during the SSM/I years. Previous work by **Armstrong and Brodzik (2001)** also suggested inconsistencies between the SMMR and SSM/I datasets. Thus the development of long-term climate-data-record quality datasets may be influenced by difficulties related to calibration between sensors.

Many other researchers have reported on algorithms that use multiple sensors to map snow cover and SWE (e.g., **Basist et al., 1996; Armstrong and Brodzik, 1998, 2001 & 2002; Standley and Barrett, 1999; Tait et al., 2000 & 2001; Kelly, 2001; Hall et al., 2002b**), and in different land covers (**Hall et al., 2001**). Visible/near-infrared data, with good spatial resolution, provide excellent snow-covered area determination under cloud-free conditions, while passive-microwave data provide all-weather, day/night snow maps, but with coarser resolution. In addition, problems arise in wet snow and thin, dry snow using passive-microwave data alone. **Hall et al., (2002b)** compared MODIS-derived maps with SSM/I-derived maps and found that the SSM/I maps showed less snow cover, particularly in the fall months, than did the MODIS maps, corroborating earlier work by **Basist et al., (1996)** and **Armstrong and Brodzik (1998 & 2002)** who compared NOAA snow maps with passive-microwave snow maps. As the snowpack wetness decreases with decreasing average air temperatures, the agreement between visible/near-infrared and passive-microwave-derived snow maps improves in January and February.

Influence of snow crystal size and shape on microwave emission. To derive the SWE using passive-microwave data, a radiative transfer approach is used in which an average crystal size of 0.3 mm (radius), a density of 300 kg m^{-3} and a spherical shape is assumed. It is also assumed that the crystals scatter radiation incoherently and independently of the path length between scattering centers. These quantities are then used in radiative transfer equations to solve the energy transfer through the snowpack (**Chang et al., 1976 & 1987**). Equations 2 and 3 are the result of this work and reasonable results are obtained from its implementation. However, if the crystal radii and snow density differ significantly from the averages and assumptions, then poor SWE values will result. Thus, current efforts are aimed at improving the methods to estimate SWE by incorporating more dynamic parameterizations of these variables.

Snow crystals consist of a myriad of shapes and sizes. Once on the ground, crystal edges or branches are quickly worn off, and the shapes become more and more rounded during destructive metamorphism. The process of freezing and thawing further rounds the crystals. Thus, the spherical shape used in the radiative transfer approximations is representative of what is observed in the field. Of course, when viewed with an electron microscope the detail is so great that individual crystals (Figure 1), can be assigned a specific shape, but the variation between even adjacent crystals can be substantial (**Rango et al., 1996; Wergin et al., 2002**). While crystal size and effective crystal size (**Mätzler, 1997**) are strongly related to microwave brightness temperature, it appears from modeling results that the shape of the snow crystal is of little consequence in accounting for the transfer of microwave radiation (at least at 0.81 cm) from the ground through the snowpack (**Foster et al., 1999 & 2000; Tsang et al., 2000**).

Active-Microwave Snow Mapping. The backscatter received by a SAR antenna is the sum of surface scattering at the air/snow interface, volume scattering within the snowpack, scattering at the snow/soil interface and volumetric scattering from the underlying surface (if applicable). Most techniques developed for mapping snow cover using synthetic-aperture radar (SAR) data show promise for mapping only wet snow (see, for example, **Rott and Nagler, 1993; Shi et al., 1994**). This is because it is difficult to distinguish dry snow from bare ground using SAR data at the X-band and lower frequencies. Volume scattering from a shallow, dry snow cover (SWE <20 cm) is undetectable at C-band (5.3 GHz, 5.6 cm), for example, because the backscatter is dominated by soil/snow scattering. Volume scattering in dry snow results from scattering at dielectric discontinuities created by the differences in electrical properties of ice crystals and air, and by ice lenses and layers. Atmospheric scattering is usually very small and can be neglected (**Ulaby and Stiles, 1980; Leconte et al., 1990; Leconte, 1995**).

In the case of wet snow (**Stiles and Ulaby, 1980; Ulaby and Stiles, 1980; Rott, 1984; Ulaby et al., 1986**), when at least one layer of the snowpack (within the penetration depth of the radar signal) becomes wet (4-5 percent liquid water content), the penetration depth of the radar signal is reduced to about 3-4 cm (or one wavelength at X-band) (**Mätzler and Schanda, 1984**). Thus there may be high contrast between snow-free ground and ground covered with wet snow thus making it possible to distinguish wet and dry land or snow when imaged with C-band radar.

In a study of wet snow using C-, L- (1.25 GHz) and P-band (440 MHz) polarimetric SAR of a mountainous area in Austria (Otztal), **Rott et al. (1992)** showed the importance of surface roughness at C- and L-band frequencies, and the increasing importance of the snow volume contribution with the longer wavelength P-band sensor. Radar polarimetry allows simultaneous measurement of the radar backscatter from a given surface at a number of different polarizations. Using European Remote Sensing Satellite (ERS)-1 images acquired before, during and after the melt period, **Koskinen et al. (1997)** successfully mapped wet snow with C-band SAR in unforested and sparsely-forested regions in northern Finland.

Rott and Nagler (1995) developed a threshold-based algorithm to map the extent of melting snow areas in mountainous regions on glaciers using ERS-1 SAR data. The classification is based on the ratio of the backscatter of the image with wet snow cover versus the backscatter of the reference image. They determined a threshold value of -3 dB by comparison with field observations. Using a single-layer backscatter model, seasonal variations as well as day/night changes in snow-covered alpine areas are shown by **Nagler and Rott (2000)** to be largely due to changes of the liquid-water content of the snowpack and the snow surface roughness. They showed the same threshold of -3dB for identifying wet snow using C-band HH Radarsat SAR and C-band ERS SAR data. Comparison of SAR-derived snow maps with Landsat-derived snow maps showed generally good results, but with a systematic underestimation of the snow extent at the edges of the snowpack using the SAR data.

In alpine regions, **Shi et al. (1994)** used single-pass SAR imagery with polarization to map wet snow cover, finding it difficult to distinguish dry snow cover from bare ground and short vegetation. Further work using single-pass, multifrequency (SIR-C/X-SAR) data (**Shi and Dozier, 1995 & 1997**) showed that the coherence between two passes provides a useful measurement and allows development of algorithms to map both wet and dry snow.

Extensive aircraft and ground measurements were obtained by the Canada Centre for Remote Sensing (CCRS) over agricultural areas in southern Québec, Canada, in 1988-1990 (**Bernier and Fortin, 1998**). It was concluded that, at C-band, volume scattering from a shallow dry snowpack (SWE < 20 cm), is not detectable. The backscatter using a C-band SAR is emanating from the snow/soil interface when the snowpack is dry. Earlier, **Mätzler and Schanda (1984)** and **Mätzler (1987)** had concluded that the backscatter change between an unfrozen bare soil and a dry snow cover over unfrozen soil was small - on the order of 1.5 dB at X-band. **Rott and Mätzler (1987)** found no significant difference between snow-free and dry-snow-covered regions at 10.4 GHz. However, there may be a potential for detecting shallow, dry snow cover with C-band SAR data when the soil beneath the snow is frozen (**Bernier and Fortin, 1998**).

Though the main scattering contribution from a dry snow cover is from the ground/snow interface, small changes in the snow can be detected using tandem pairs from repeat-pass interferometric SAR (InSAR) data. Using the coherence measurement of repeat passes, **Shi and Dozier (2000 a & b)** found that both wet and dry snow can be mapped as evidenced by comparison of snow mapped using Landsat imagery. Refraction of microwaves in dry snow was shown to have a significant effect on the interferometric phase difference and a relationship between changes in SWE and the interferometric phase was derived. Using three tandem pairs of InSAR data, **Guneriussen et al. (2001)** found that a snow density of 0.2 g/cm^2 at 23° incidence angle gives phase wrapping for changes in snow depth of 16.4 cm and equals a SWE of 3.3 cm. They also concluded that even light snowfall and/or small changes in snow properties between InSAR images introduces a significant height error in digital-elevation models (DEMs) derived from snow-covered glaciers and ice sheets.

There are conflicting results showing the sensitivity of radars to snow water equivalent. **Ulaby et al. (1977)** found that radar sensitivity to total snow water equivalent (for wet snow) increased in magnitude with increasing frequency (or shorter wavelength) and is almost independent of angle for angles of incidence $>30^\circ$, particularly at higher frequencies. **Goodison et al. (1980)** found little or no relationship between radar return and snowpack properties under either wet or dry snow conditions using L-band airborne SAR. A uniform, low return was found for a given area under both snow-free and snow-covered conditions. However, using X-band, in their study area near Ottawa, Ontario, non-forested areas exhibited higher backscatter when snow cover was present. Areas of ice and dense snow were observed to produce relatively higher radar returns using the X-band SAR (**Goodison et al., 1980**). More recently, **Shi and Dozier (2000a)** developed an algorithm for estimating snow density using L-band VV and HH measurements based on numerically-simulated backscatter coefficients. Comparison of the estimated snow

density from three Shuttle Imaging Radar-C (SIR-C)/X-band images with field snow density measurements shows an absolute RMSE of 42 kg m^{-3} and a relative 13% error. **Shi and Dozier (2000b)** also estimated snow depth and particle size using SIR-C/X-SAR imagery from a physically-based first order backscattering model. Using Radarsat SAR data, **Bernier et al. (1999)** established a relationship between the backscattering ratios of a winter (snow covered) image and a fall (snow-free) image to estimate the snowpack thermal resistance. They then estimate the SWE from the thermal resistance and the measured mean density. In related work, **Gauthier et al. (2001)** used Radarsat C-band ScanSAR data to derive SWE in the La Grande River basin in northern Québec where they found ScanSAR-derived SWE values to be similar ($\pm 12\%$) to those derived from in-situ snow measurements in January and March 1999.

A promising technology for measuring snow cover is scatterometry. A Ku-band (14.6 GHz) scatterometer operated for three months in July – September 1978 on the Seasat satellite, and results showed that some of the glacier facies could be mapped using derived backscatter images (**Long and Drinkwater, 1994**). The NASA Scatterometer (NSCAT), operated on the Advanced Environmental Observation Satellite (ADEOS) from September 1996 to June 1997 and also permitted study of melt zones on Greenland (**Long and Drinkwater, 1999**). Timing of melt onset was detected by **Nghiem et al. (2001)** on the Greenland Ice Sheet, by a sharp decrease in backscatter, and verified with in-situ measurements, using the SeaWinds scatterometer on the QuikSCAT satellite. For seasonal snow cover, **Nghiem and Tsai (2001)** show that NSCAT backscatter patterns reveal boundaries that correspond to various snow classes as defined by **Sturm et al. (1995)**. Additionally they show rapid changes in the backscatter over the northern plains of the United States and the Canadian prairies that led to the major spring 1997 floods in the mid-western United States and southern Canada (Figure 7).

Conclusion

Satellite remote sensing has been used to map snow cover for nearly 40 years. Decade-scale climate-data-record-quality records of snow-covered area are already in existence for the Northern Hemisphere (**Robinson and Frei, 2000**) plus are useful in climate models, and are invaluable for other climate studies. We can now extend the snow-covered area record using SMMR, SSM/I MODIS and AMSR data, to the global scale, and provide climate-data-quality maps of not only snow-covered area, but also SWE and snow albedo using visible, near infrared, and passive-microwave sensors such as the MODIS and AMSR.

The trend has been toward increasingly automatically-processed quantitative maps with error bars provided. Automated processing is necessary so that consistent products can be derived from the observations and long duration data sets might ultimately be available for long-term water-cycle studies. The error estimates associated with snow products are also essential if the products are to be used effectively in combination with catchment, land surface or climate models. This is because models that require snow state parameters often require the errors associated with the estimated snow states, especially if

data assimilation techniques are used to generate blended products. Whether the snow products are used for initial conditioning or as a forcing variable in a model, or whether the products are used in their own right, the role of remotely sensed observations of snow will continue to be important and are set to play an increasingly important role in climate and hydrological forecasting.

Future sensors will permit automated algorithms to be used to create maps that are consistent with existing maps so that the confidence level of the long-term (~40 year) record is high. The quality allows them to be amenable to comparison with long-term records of other geophysical parameters such as global sea ice, and for input to general-circulation models.

Acknowledgments

The authors would like to thank Dr. Andrew G. Klein of Texas A & M University, for helpful comments on parts of the manuscript.

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Figure captions

Figure 1. This image of a newly-fallen hexagonal plate snow crystal was taken with an Electron Scanning Microscope and shows a classic dendritic snow crystal, having a central hexagonal plate, but lacking sharp edges. Because of the air temperature at the time of collection (close to 0°C), this crystal may have undergone some sublimation (Wergin et al., 1996).

Figure 2. MODIS 8-day composite climate-modeling grid (CMG) global snow-cover map.

Figure 3. Illustration of the effect of different snow crystals on snow reflectance (from Choudhury and Chang, 1979).

Figure 4. Prototype of snow albedo product - north central United States and southern Canada - February 16, 2001. (Courtesy of Andrew Klein, Texas A & M University, College Station, TX.)

Figure 5. Estimated snow depth for the Northern Hemisphere from AMSR-E data for 7 December 2002 (from R.E.J. Kelly and A.T.C. Chang, unpublished data).

Figure 6. Snow water equivalent (in mm) over the Canadian Prairie region, derived from DMSP SSM/I data for 15 February 2003. Areas of highest snow water equivalent generally correspond to the areas where the snow cover is deepest. (Courtesy of the Climate Research Branch, Meteorological Service of Canada, Environment Canada.)

Figure 7. NSCAT backscatter signatures over snow cover corresponding to snow events leading to the 1997 flood in the northern plains of the United States and the Canadian prairies: a) period of snowmelt from March 19 to March 25, 1997; b) period of the blizzard from April 2 to April 8, 1997; and c) period of rapid snow retreat from April 16 to April 22, 1997 (Nghiem et al., 2001). (©2001 IEEE) (Courtesy of Son Nghiem, Jet Propulsion Laboratory, Pasadena, CA).

Table

Table 1. Comparison of characteristics of passive-microwave sensors (**from Foster et al., in preparation**).

Table 1. Comparison of characteristics of passive-microwave sensors (from Foster et al., in preparation).

	SMMR	SSM/I	AMSR-E
Platform	Nimbus-7	DMSP F-8, 11, 13	Aqua
Period of Operation	1979-87	1987-present	2002-present
Data Acquisition	every other day	daily	daily
Swath Width	780 km	1400 km	1600 km
Frequency (GHz)	18.0 37.0	19.35 37.0	18.7 36.5
Spatial Resolution (km)	60x40 (18 GHz)	69x43 (19.4 GHz)	28x16 (19.7 GHz)
	30x20 (37 GHz)	37x29 (37 GHz)	14x8 (36.5 GHz)
Polarization	H & V	H & V	H & V
Orbital Timing (Eq. Crossing) (for minimum temperature)	midnight	6:00 a.m.	1:30 a.m.